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# Mantle plumes and mantle dynamics in the Wilson cycle

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**Abstract:** This review discusses the thermal evolution of the mantle following large-scale tectonic activities such as continental collision and continental rifting. About 300 myr ago, continental material amalgamated through the large-scale subduction of oceanic seafloor, marking the termination of one or more oceanic basins (e.g. Wilson cycles) and the formation of the supercontinent Pangaea. The present day location of the continents is due to the rifting apart of Pangaea, with the dispersal of the supercontinent being characterized by increased volcanic activity linked to the generation of deep mantle plumes. The discussion presented here investigates theories regarding the thermal evolution of the mantle (e.g. mantle temperatures and sub-continental plumes) following the formation of a supercontinent. Rifting, orogenesis and mass eruptions from large igneous provinces change the landscape of the lithosphere, whereas processes related to the initiation and termination of oceanic subduction have a profound impact on deep mantle reservoirs and thermal upwelling through the modification of mantle flow. Upwelling and downwelling in mantle convection are dynamically linked and can influence processes from the crust to the core, placing the Wilson cycle and the evolution of oceans at the forefront of our dynamic Earth.

The theory of plate tectonics describes the movement of the Earth's lithosphere, while the convective motion of the Earth's mantle drives the tectonic plates to determine the present day position of the continents. Geological features formed by continent–continent collisions (e.g. mountain ranges, faulting) indicate that the North American continent consists of 13 major cratons (old and stable continental lithosphere) amalgamated by plate tectonics (Hoffman 1988). Furthermore, similar fossils, flora and fauna on the land masses on either side of the Atlantic Ocean indicate that North and South America were once attached to the African and European continents (Wilson 1966). These descriptions of the movement of the continents are a corollary of the theory of plate tectonics. However, the dynamic processes involved in plate tectonic motion, and its relation to the thermal evolution of the mantle, are still being debated.

In 1963, at the University of Toronto, John Tuzo Wilson added a pivotal concept to the then peripheral theory of plate tectonics. Wilson (1963) suggested that the Hawaiian island volcanoes were created by the NW shifting of the Pacific tectonic plate over a fixed mantle hotspot. Plate tectonic theory began to generate more interest throughout the 1960s and previous work in support of the hypothesis was brought to the forefront of earth science research (e.g. Agrand 1924; Wegener 1924; Holmes 1931; Du Toit 1937). In 1966, based on evidence in the fossil record and the dating of vestiges of ancient volcanoes, Wilson

proposed a cycle describing the opening and closing of oceanic basins, and therefore a method of amalgamating continental material that would subsequently be dispersed. Wilson (1966), building on previous studies (e.g. Hess 1962; Vine & Matthews 1963; Wilson 1965), outlined a cycle of ocean basin evolution: the dispersal (or rifting) of a continent; continental drift, seafloor spreading and the formation of oceanic basins; new subduction initiation and the subsequent closure of oceanic basins through oceanic lithosphere subduction; and continent–continent collision and closure of the oceanic basin. This lifecycle of oceans was later termed the Wilson cycle by Dewey & Burke (1974). The aggregation and dispersal of continents into a supercontinent (e.g. the supercontinent cycle; Worsley *et al.* 1982, 1984; Nance *et al.* 1988; Rogers & Santosh 2003; Nance & Murphy 2013; Matthews *et al.* 2016) is intrinsically linked to the Wilson cycle through oceanic closures. At present, the general form of the supercontinent cycle consists of four parts.

- (1) Continental material aggregates over a large downwelling in the mantle to form a supercontinent (Zhong *et al.* 2007).
- (2) The formation of a supercontinent is characterized by subduction on its margins (Li *et al.* 2008; Li & Zhong 2009), with large-scale subduction into the mantle.
- (3) Thermal insulation by the continent traps the underlying heat and the repositioning of

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subduction zones focuses thermal anomalies below the supercontinent. A plume is formed beneath the supercontinent 50–100 myr after continental accretion (Li *et al.* 2003).

- (4) The supercontinent breaks up along pre-existing suture zones (Butler & Jarvis 2004; Murphy *et al.* 2006, 2008) as a result of divergent horizontal flow and the lithosphere's tensional yield stress being exceeded. The timescale for the full cycle to be repeated is c. 500–700 myr (e.g. Nance *et al.* 1988; Rogers & Santosh 2003; Li & Zhong 2009; Nance & Murphy 2013; Evans *et al.* 2016; Matthews *et al.* 2016; Green *et al.* 2018).

Figure 1 shows a cartoon of the basic supercontinent cycle. Geologists and geophysicists have progressed the theory of plate tectonics over the past 50 years. However, the mechanisms involved in supercontinent formation and dispersal are still divisive. Despite recent advances in our understanding of mantle convection, the roles of circum-supercontinent subduction (Li *et al.* 2008) (step 2, Fig. 1) and continental thermal insulation (step 3, Fig. 1) in the generation of sub-continental plumes (step 4, Fig. 1) remain unclear.

This review outlines the features of mantle dynamics following cyclic plate tectonic processes. Supercontinent dispersal can produce several internal oceans, forming a number of separate Wilson cycles (Green *et al.* 2018). Supercontinent formation may require that more than one lifecycle of an ocean (e.g. a Wilson cycle) is terminated (e.g. Burke 2011). Over the duration of the formation of a supercontinent, which may have a timescale of hundreds of million years (Hoffman 1991; Torsvik 2003; Li *et al.* 2008), a number of oceans may close. This indicates that the supercontinent cycle would have a greater impact on the thermal evolution of the mantle than a simple, singular Wilson cycle. As a result, this review highlights ideas relating to supercontinent formation and dispersal to outline a broad, global view of dynamics – with an application to the processes involved in the Wilson cycle.

Following this discussion of the supercontinent cycle, this review then outlines mantle dynamics in relation to continent collision, thermal insulation, plume formation and large-scale oceanic subduction, ending with a discussion of the relation between surface tectonics and deep mantle processes.

## Continent collision: supercontinent formation

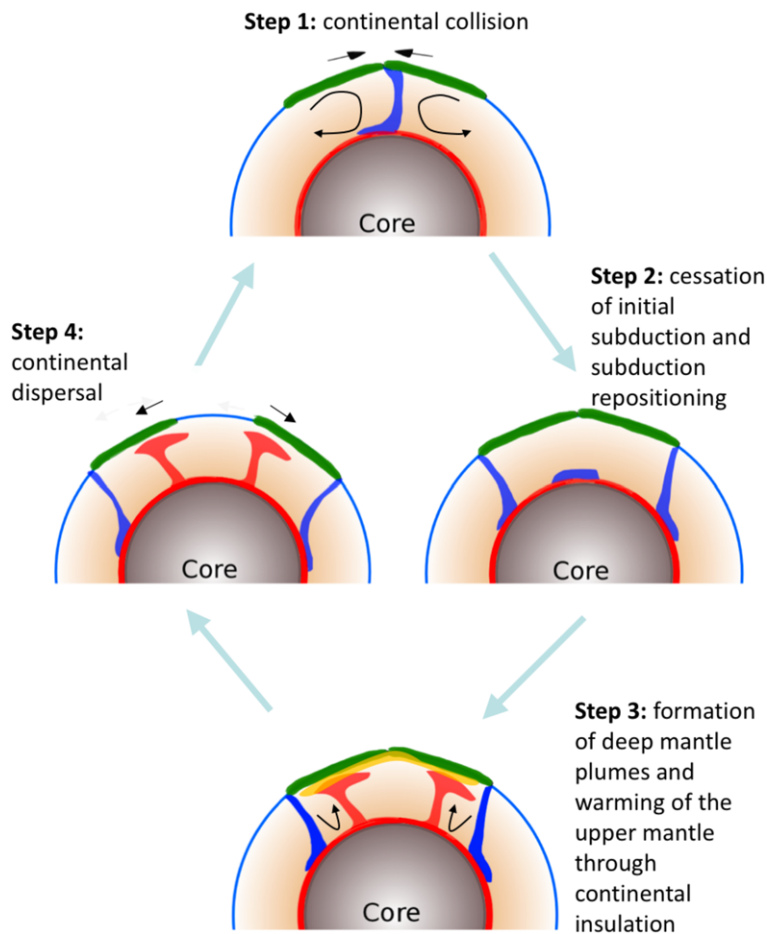
Plate movement reconstructions (using palaeomagnetism) and geological analyses of orogenesis (i.e. mountain-building) hold information pertaining to the supercontinent cycle. Studies analysing the

timing of continent–continent collisions and rifting sequences show that the land masses of Gondwana (the African, Antarctic, India, Australian and South American plates) and Laurasia (the Eurasian and North American plates) formed the supercontinent Pangaea near the equator at c. 320 Ma (Smith *et al.* 1981; Hoffman 1991; Scotese 2001). The Appalachian and Ural mountain belts were generated as a result of this collision. Global plate reconstructions and analyses of volcanic arc lavas show that Pangaea was ringed by subduction during the lifespan of the fully assembled supercontinent (Fig. 2) (e.g. Li *et al.* 2008; Li & Zhong 2009; Matthews *et al.* 2016). The break-up of the supercontinent Pangaea is thought to have occurred in a number of stages: North America separating from the land mass at c. 180 Ma (starting the opening of the North Atlantic Ocean), followed by the dispersal of the Antarctic–Australian, Indian and South American continents between 140 and 100 Ma (Smith *et al.* 1981; Hoffman 1991; Scotese 2001), with the final separation of Australia and Antarctica occurring in the Paleocene (Veevers & McElhinny 1976).

The supercontinent Rodinia formed with Laurentia (the North American craton) at its centre and generated (among other mountain belts) the Grenville Orogeny (including the Laurentian mountain range of Quebec) (Hoffman 1991; Dalziel 1991; Moores 1991; Torsvik 2003; Li *et al.* 2004, 2008). Rodinia was fully assembled by c. 900 Ma and subduction featured on its margins, similar to Pangaea (Li *et al.* 2008). Rodinia's break-up occurred between 720 and 650 Ma (Li *et al.* 2008, 2013; Li & Evans 2011), starting with South China separating from Laurentia at c. 750 Ma (Li *et al.* 2008).

The formation of Pangaea and Rodinia has been attributed to large-scale mantle downwellings amassing continental material (e.g. Scotese 2001; Murphy and Nance, 2003; Zhong *et al.* 2007; Murphy *et al.* 2009). In the Wilson cycle, the closing of oceanic basins due to subduction has been linked to orogenesis and continental growth (e.g. Wilson 1966; Dewey 1969). However, in the supercontinent cycle, two end-member modes of subduction may play a part (Fig. 3). The geological and Sm/Nd isotopic record suggests that supercontinents may form via introversion, in which oceans that are interior to the continental material are preferentially subducted, or extroversion, in which exterior ocean floor is preferentially subducted (Murphy *et al.* 2009). The process of introversion (Fig. 3a) has been shown to have occurred in the formation of Pangaea (e.g. Murphy and Nance, 2003). The closing of the Iapetus and Rheic oceans (the latter through a sudden reversal in oceanic plate motion) are believed to be fundamental in the introversion method of amalgamating the supercontinent Pangaea (Murphy *et al.* 2006; Nance *et al.* 2012).

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**Fig. 1.** A cartoon of a simple supercontinent, the starting point for this review. Step 1, a supercontinent is amassed through a super-downwelling. Step 2, subduction then forms on the margins of the continent, generating sub-continental plumes due to mantle return flow and warming of the mantle through continental insulation (step 3). Step 4, the continental plumes facilitate the dispersal of the supercontinent.

After analysing the topology within supercontinents, [Santosh \*et al.\* \(2009\)](#) proposed the large-scale downwellings that amass continental material to be produced at a Y-shaped plate boundary junction. This configuration would promote stronger downwellings that could generate the runaway subduction of oceanic material ([Santosh \*et al.\* 2009](#)). As a result, subduction plays a key part in determining the location and configuration of a future supercontinent. If introversion processes are dominant ([Fig. 3a](#)), then the Atlantic Ocean will act as the present day version of the Iapetus and Rheic oceans in the future formation of Amasia ([Hoffman 1997](#)). If extroversion processes are dominant ([Fig. 3b](#)), then the Pacific will close to form the Novopangaea supercontinent ([Nield 2007](#)). The simultaneous closure of the Atlantic and Pacific oceans would form the supercontinent

Aurica, as described by [Duarte \*et al.\* \(2018\)](#). Another example is the northwards migration of the North American and Eurasian plates to produce a supercontinent through the closure of the Arctic Ocean ([Hoffman 1997](#); [Mitchell \*et al.\* 2012](#)). The closure of the Arctic Ocean has been categorized as orthoversion ([Fig. 3c](#)), a third subduction option where a supercontinent forms orthogonal to the centroid of the previous supercontinent ([Evans 2003](#); [Mitchell \*et al.\* 2012](#)).

### Thermal insulation due to supercontinent formation

As a result of its relatively greater buoyancy, continental material remains on the Earth's surface while

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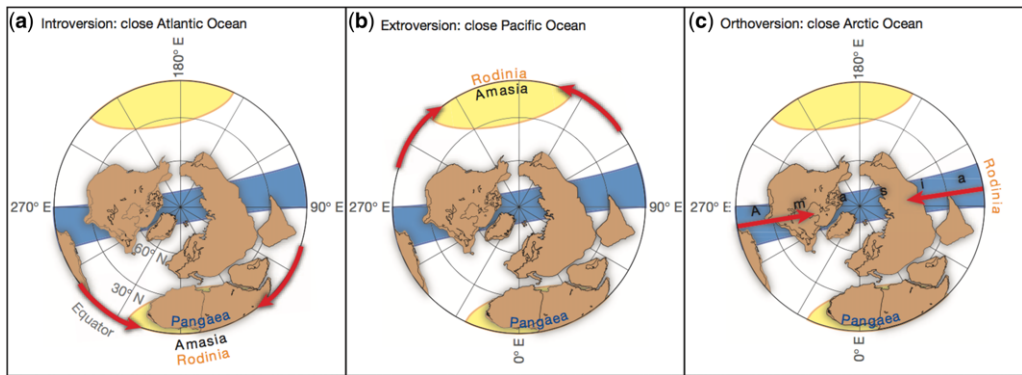
**Fig. 2.** Circum-supercontinent subduction. Reconstructions of the supercontinent Pangaea during the late Permian. Reprinted from [Torsvik \(2003\)](#) with permission.

oceanic plates are subducted. Continental lithosphere inhibits heat loss from the Earth's interior, relative to oceanic lithosphere, due to its greater thickness. [Anderson \(1982\)](#) first suggested that continental insulation could control the supercontinent cycle by having a dramatic effect on the underlying mantle. Over a long timescale, the supercontinent would trap excess heat and cause uplift (through thermal expansion), partial melting of the mantle and, ultimately, the dispersal of continental material ([Anderson 1982](#)). A large geoid high would be generated below the supercontinent, similar to the present day geoid profile over Africa, through the thermal expansion caused by continental insulation. [Gurnis \(1988\)](#) produced the first numerical models of supercontinent formation and dispersal and found that a large continent could inhibit mantle cooling, leading to fragmentation of the continent through mantle

overheating. Looking back over Earth's history for data on such a geodynamic event, a study compiling thermal gradients and the ages of metamorphism from the Eoarchean to Cenozoic found a thermal maximum in the mid-Mesoproterozoic, which was hypothesized to be a result of thermal insulation of the mantle beneath the continental lithosphere of the supercontinent Nuna ([Brown & Johnson 2018](#)).

Many numerical studies have shown that the combination of supercontinent coverage and insulation can generate sub-supercontinental temperatures higher than sub-oceanic mantle material, suggesting that continental insulation acts as the main driver for supercontinent break-up (e.g. [Gurnis 1988](#); [Zhong & Gurnis 1993](#); [Lowman & Jarvis 1993, 1999](#); [Bobrov et al. 1999](#); [Yoshida et al. 1999](#); [Phillips & Bunge 2005](#); [Coltice et al. 2007](#); [Trubitsyn et al. 2008](#); [Coltice et al. 2009](#); [Phillips et al. 2009](#); [Phillips &](#)

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**Fig. 3.** Predicted locations of a future supercontinent (here named Amasia) according to three possible models of the supercontinent cycle: (a) introversion; (b) extroversion and (c) orthoversion. Image reprinted from Mitchell *et al.* (2012) with permission. Red arrows indicate where the ocean basins would close according to each model. Continents are shown in present day coordinates. Yellow equatorial circles represent areas of deep mantle heterogeneities and supercontinent-induced mantle upwellings. The labelled centres of Pangaea and Rodinia are the conjectured locations of each supercontinent's minimum moment of inertia; see Mitchell *et al.* (2012) for details.

Coltice 2010; Yoshida 2010; Rolf *et al.* 2012). A geochemical study into ancient lava samples from the Atlantic Ocean indicates increased mantle temperatures relative to Pacific Ocean samples during the dispersal of the supercontinent Pangaea (Brandl *et al.* 2013). By analysing lava samples from the past 170 myr, Brandl *et al.* (2013) showed that the post-supercontinent upper mantle beneath the Atlantic Ocean was 150 K warmer than the present day values. By comparing mid-ocean ridges in the Atlantic and the Pacific (where the latter has samples that formed >2000 km from the nearest continental craton), Brandl *et al.* (2013) found that upper mantle temperatures in the Atlantic Ocean remained high for 60–70 myr, before returning to temperatures like those found beneath the Pacific Ocean. They attribute the temperature difference to continental insulation by the supercontinent Pangaea.

### Large igneous provinces: link to supercontinent formation and destruction

Past deep mantle plumes are thought to be manifested on the Earth's surface by expansive areas of igneous material, erupted over relatively short geological timescales – for example, large igneous provinces (LIPs) (Burke & Torsvik 2004). Although the origin of LIPs remains controversial (White & McKenzie 1989; Coffin & Eldholm 1994; Ernst *et al.* 2005; Foulger 2007; Campbell & Kerr 2007), a deep mantle source can be inferred from regional domal uplift due to the sub-crustal arrival of buoyant plume material (e.g. Sleep 1990; Davies 1999; Sengör 2001; Rainbird & Ernst 2001; Saunders *et al.* 2007) and possibly through tomographic images of present day hotspots

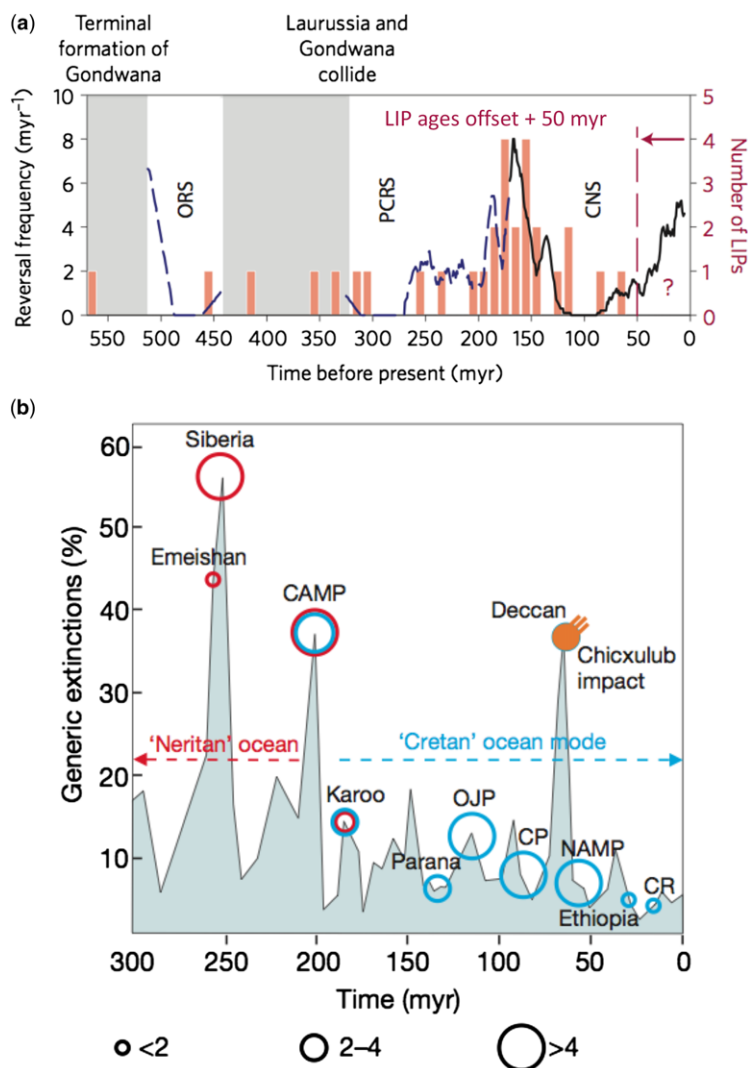
(French & Romanowicz 2015). Analysing the rock record over Earth's history shows little LIP activity during the amalgamation stage of the supercontinent cycle (e.g. Yale & Carpenter 1998; Ernst *et al.* 2005; Ernst & Bleeker 2010). However, after a supercontinent has been formed for a period of time, the number of LIPs increases on a global scale (e.g. Yale & Carpenter 1998; Ernst *et al.* 2005; Ernst & Bleeker 2010; Sobolev *et al.* 2011), as shown in Figure 4a. LIPs have a large-scale impact on the surface and represent a significant mantle event, producing a thinned lithosphere (White & McKenzie 1989; Garfunkel 2008) that could generate continental break-up (Burke & Dewey 1973; White & McKenzie 1989; Hill 1991; Courtillot *et al.* 1999) as well as mass extinction events (Wignall 2001; Courtillot & Renne 2003; Sobolev *et al.* 2011) (Fig. 4b). There is also evidence to suggest a link between geomagnetic variations and mantle convection. Simulations of the geodynamo suggest that transitions from periods of rapid polarity reversals to periods of prolonged stability may have been triggered by a decrease in core–mantle boundary heat flow, either globally or in equatorial regions (e.g. Larson & Olson 1991; Biggin *et al.* 2012). This decrease in the core–mantle boundary heat flow could be related to the formation of a supercontinent, but before the repositioning of subduction and the generation of plumes (Larson & Olson 1991) (Fig. 1).

### Mantle return flow and the formation of plumes

Zhong *et al.* (2007) presented two planform regimes for the Earth in three-dimensional spherical shell



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**Fig. 4.** (a) Average geomagnetic reversal frequency and eruption ages (Torsvik *et al.* 2010) of large igneous provinces (LIPs; offset by +50 myr) that have not yet been subducted. Mantle plume heads leaving the core–mantle boundary (CMB) may reflect enhanced heat flow out of the core, potentially increasing the reversal frequency tens of millions of years before the resulting eruption of the LIP. Allowing for an average rise time of 50 myr results in a broad correlation that would associate geomagnetic reversal hyperactivity in the Mid Jurassic with widespread LIP emplacement in the Mid Cretaceous. In the period 0–50 myr, mantle plume heads that had left the CMB would not yet have reached the surface. CNS, Cretaceous Normal Superchron; ORS, Ordovician Reversed Superchron; PCRS, Permo-Carboniferous Reversed Superchron. Figure reprinted from Biggin *et al.* (2012) with permission. (b) Plot of mass extinction intensity (light blue field) with major LIPs (circles) against geological time (modified from White & Saunders (2005)), together with the timing of different ocean modes (Ridgwell 2005). Circle colours denote the timing of LIPs relative to ocean modes: blue, Cretan mode; red Neritan mode; blue and red together, transition mode (see Sobolev *et al.* (2011) for details). The scale of the circle sizes is in millions of cubic kilometres. CAMP, Central Atlantic Magmatic Province; CP, Caribbean Plateaux; CR, Columbian River basalts; NAMP, Northern Atlantic Magmatic Provinces, OJP, Ontong Java. Reprinted from Sobolev *et al.* (2011) with permission.

mantle convection models with mobile lids. When a supercontinent is absent, the mantle planform is characterized by a spherical harmonic degree-1

structure, with a major upwelling in one hemisphere and a major downwelling in the other (Zhong *et al.* 2007). Following the placement of a supercontinent

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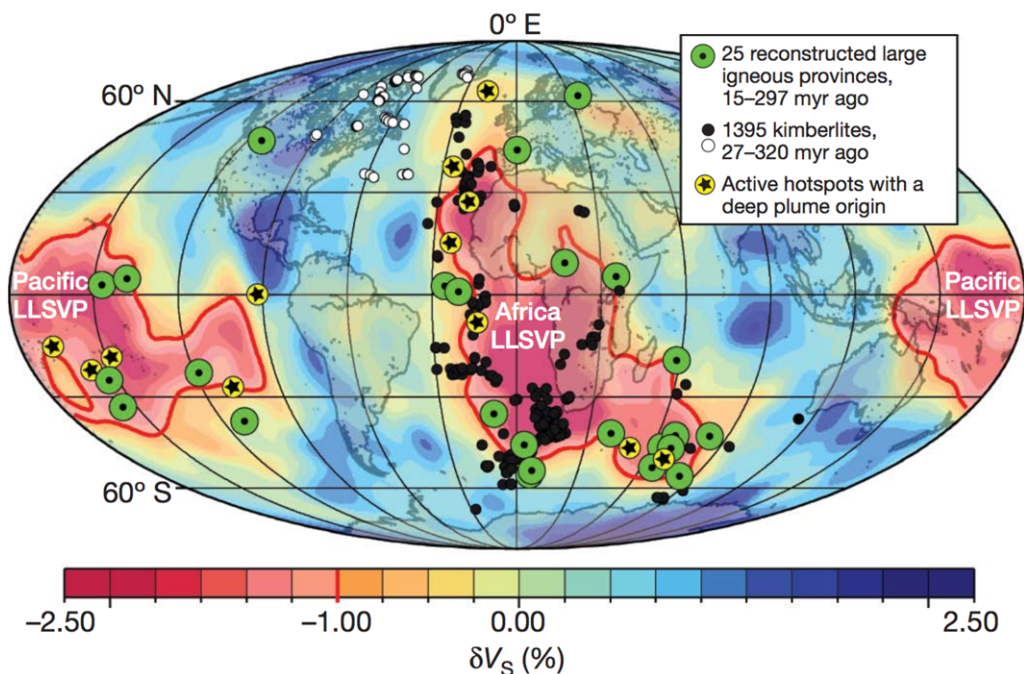
above the downwelling, a dominantly degree-2 planform develops with two antipodal major upwellings. Consequently, a dominantly degree-1 planform acts to form a supercontinent which, once fully assembled, changes the mantle planform to degree-2. The formation of a stationary supercontinent generates subduction on its edges, which, in turn, generates a super-plume *c.* 50 myr later, which subsequently facilitates the dispersal of continental material (e.g. Fig. 1). The timing of the mantle return flow, where subduction going down into the lower mantle produces upwellings returning to the upper mantle, is roughly in keeping with the generation of LIPs post-supercontinent formation (Fig. 4) and the estimated plume rise time (Olson *et al.* 1987; Thompson & Tackley 1998).

Zhong *et al.* (2007) suggest that the Africa and Pacific antipodal super-plumes (the basis of the degree-2 structure of the present day mantle) are a consequence of the supercontinent cycle, with the Pacific upwellings being dominant during the formation of Pangaea. Therefore the mantle may modulate between dominantly degree-1 and dominantly degree-2 planforms for supercontinent formation

and dispersal, respectively, and mantle plumes are generated by the formation of the supercontinent (Zhong *et al.* 2007).

### Present day mantle dynamics from seismic imaging studies

Below the surface, the thermal field of the present day mantle may also hold information pertaining to supercontinent cycle dynamics. Figure 5 shows a horizontal cross-section of a global seismic tomography model depicting the relative variations in shear velocity (with respect to the mean) at 2800 km depth in the mantle (Torsvik *et al.* 2010). Near the core–mantle boundary, anomalously warm material (characterized by slow shear wave velocities) is present beneath the Pacific and African plates, with the latter lying below the site of the last supercontinent Pangaea. As a result of these present day temperature anomalies, the mantle's thermal and geoid profiles are characterized by a degree-2 harmonic structure, in keeping with the numerical modelling of the supercontinent cycle (e.g. Zhong *et al.* 2007).

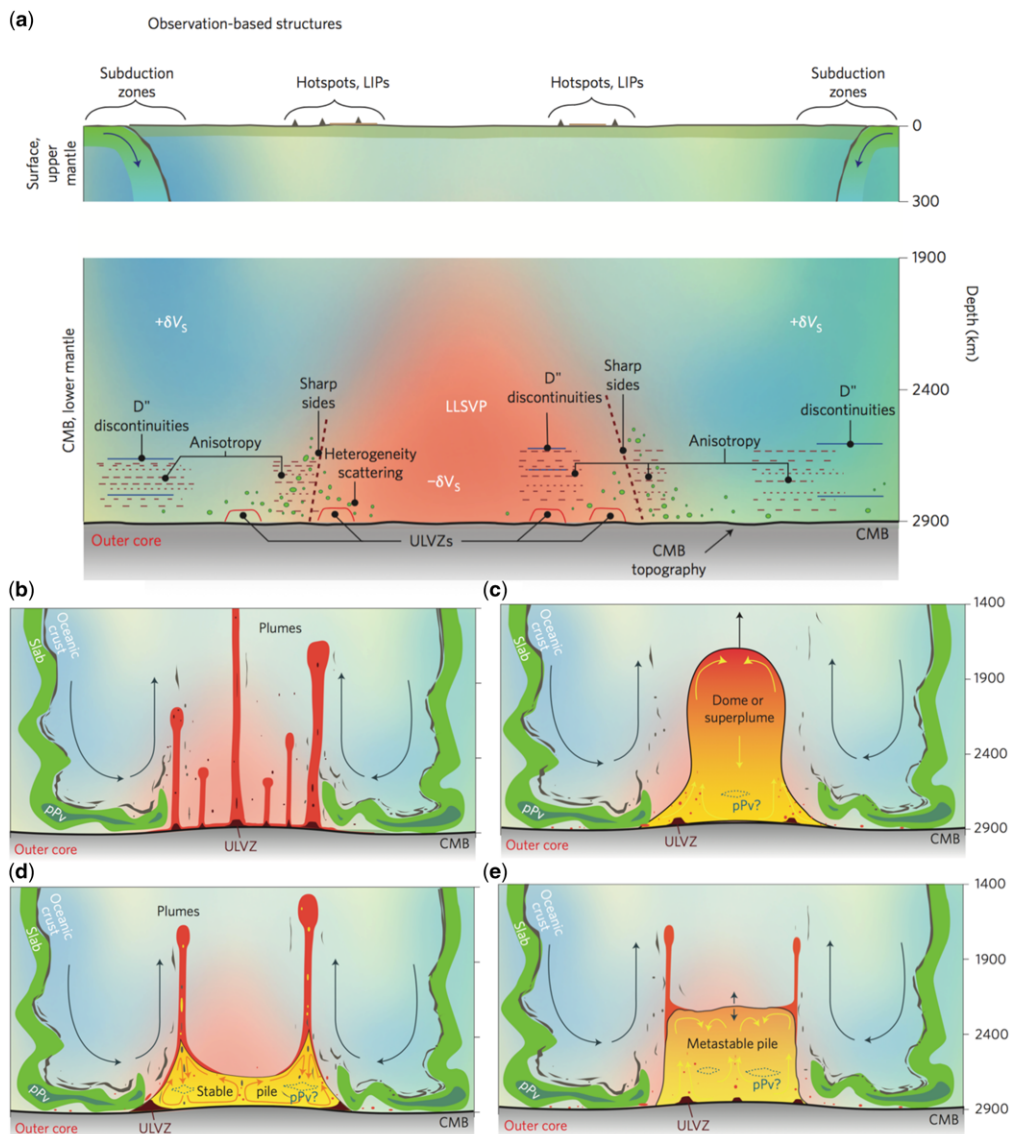


**Fig. 5.** Reconstructed large igneous provinces and kimberlites for the past 320 myr with respect to shear wave anomalies at the base of the mantle. The deep mantle (2800 km on the SMEAN tomography model (Becker & Boschi 2002)) is dominated by two large low shear velocity provinces (LLSVPs) beneath Africa and the Pacific Ocean. Kimberlite locations are given by black dots over the LLSVPs and white dots in Canada. The present day continents are shown as a background alongside hotspots that are thought to be of deep mantle origin (Montelli *et al.* 2006). Image reprinted from Torsvik *et al.* (2010) with permission.



As shown in Figure 5, the two regions are characterized by low shear wave velocities. As a result, they are widely known as large low shear velocity provinces (LLSVPs). Ultra-low velocity zones (ULVZ) also reside on the core–mantle boundary, but are an order of magnitude smaller (Garnero & Helmberger 1996; Williams & Garnero 1996; Williams

*et al.* 1998; Garnero *et al.* 2007; Garnero & McNamara 2008). At present, there is strong debate as to whether the LLSVPs and ULVZs are thermal and/or compositional features that are chemically distinct from the surrounding mantle (as outlined in McNamara 2018). Figure 6 describes some of the theories regarding the nature of the lower mantle seismic



**Fig. 6.** (a) Surface features (upper panel) and seismically determined lower mantle phenomena (lower panel). (b–e) Idealized possibilities proposed to explain large low shear velocity provinces (LLSVPs). In all cases, subducted material surrounds the structure of interest that maps as the LLSVP. (b) Plume cluster. (c) Thermochemical superplume. (d) Stable thermochemical pile. (e) Metastable thermochemical pile. LIPs, large igneous provinces; CMB, core–mantle boundary; ULVZs, ultra-low velocity zones. Reprinted from Garnero *et al.* (2016) with permission.

features (with more details featured in [Garnero \*et al.\* 2016](#)). The two large anomalies beneath Africa and the Pacific Ocean ([Fig. 5](#)) are thought to be mantle plume clusters ([Fig. 6b](#)) or super-plumes ([Fig. 6c](#)), formed through processes related to the supercontinent cycle (e.g. [Schubert \*et al.\* 2004](#); [Zhong \*et al.\* 2007](#); [Li \*et al.\* 2008](#); [Li & Zhong 2009](#)), and thermal in nature. However, the possibility of the LLSVPs being chemically distinct from the surrounding mantle also allows different geodynamic scenarios for plume generation (e.g. [Tackley 1998](#); [Nakagawa \*et al.\* 2009](#); [Bull \*et al.\* 2009](#); [Zhang \*et al.\* 2010](#); [Torsvik \*et al.\* 2010](#); [Burke 2011](#)). In particular, the relative stability of LLSVPs in their core–mantle boundary position may generate plumes from the edges of stable ([Fig. 6d](#)) or metastable thermochemical piles ([Fig. 6e](#)).

The diverse nature of interpretation of LLSVPs within the geophysics community highlights the difficulties in constraining mantle dynamics ([Torsvik \*et al.\* 2010](#); [Burke 2011](#); [Davies \*et al.\* 2012](#); [Garnero \*et al.\* 2016](#); [Zaroli \*et al.\* 2017](#)). Early studies of tomographic normal model data found that the two deep anomalies were characterized by higher than average densities ([Ishii & Tromp 1999, 2001, 2004](#); [Trampert \*et al.\* 2004](#)), whereas a study using Stoneley mode data found that LLSVP regions show lower than average densities ([Koelemeijer \*et al.\* 2017](#)). Compositional heterogeneities and/or a phase change post-perovskite could also fit seismic observations in the lowermost mantle (e.g. [Trampert \*et al.\* 2001](#); [Deschamps & Trampert 2003](#); [Trampert \*et al.\* 2004](#); [Hernlund & Houser 2008](#); [Mosca \*et al.\* 2012](#); [Koelemeijer \*et al.\* 2016](#)). The possibility that LLSVPs are purely thermal plumes has been put forward ([Schubert \*et al.\* 2004](#); [Davies & Davies 2009](#)), with numerous studies highlighting that the low tomographic resolution in the deep mantle may smear our view of the geodynamic features (e.g. [Ritsema \*et al.\* 2007](#); [Schuberth \*et al.\* 2009](#); [Bull \*et al.\* 2009, 2010](#); [Davies \*et al.\* 2012](#); [Zaroli \*et al.\* 2017](#)).

Despite the uncertainties about the composition of LLSVPs ([Labrosse \*et al.\* 2007](#); [Davies \*et al.\* 2012](#); [Deschamps \*et al.\* 2012](#); [Garnero \*et al.\* 2016](#)), what is clear is that mantle plumes originate from these LLSVP regions through one mechanism or another ([Fig. 6](#)), producing LIPs that may affect plate motion (e.g. [Van Hinsbergen \*et al.\* 2011](#)) and, in turn, subduction initiation (e.g. [Gerya \*et al.\* 2015](#)), alongside other large-scale geodynamic and environmental effects ([Fig. 4](#)).

## Discussion

Mantle convection is a dynamic process whereby subduction and plumes operate in a feedback system: plumes influence plate motion and subduction, and

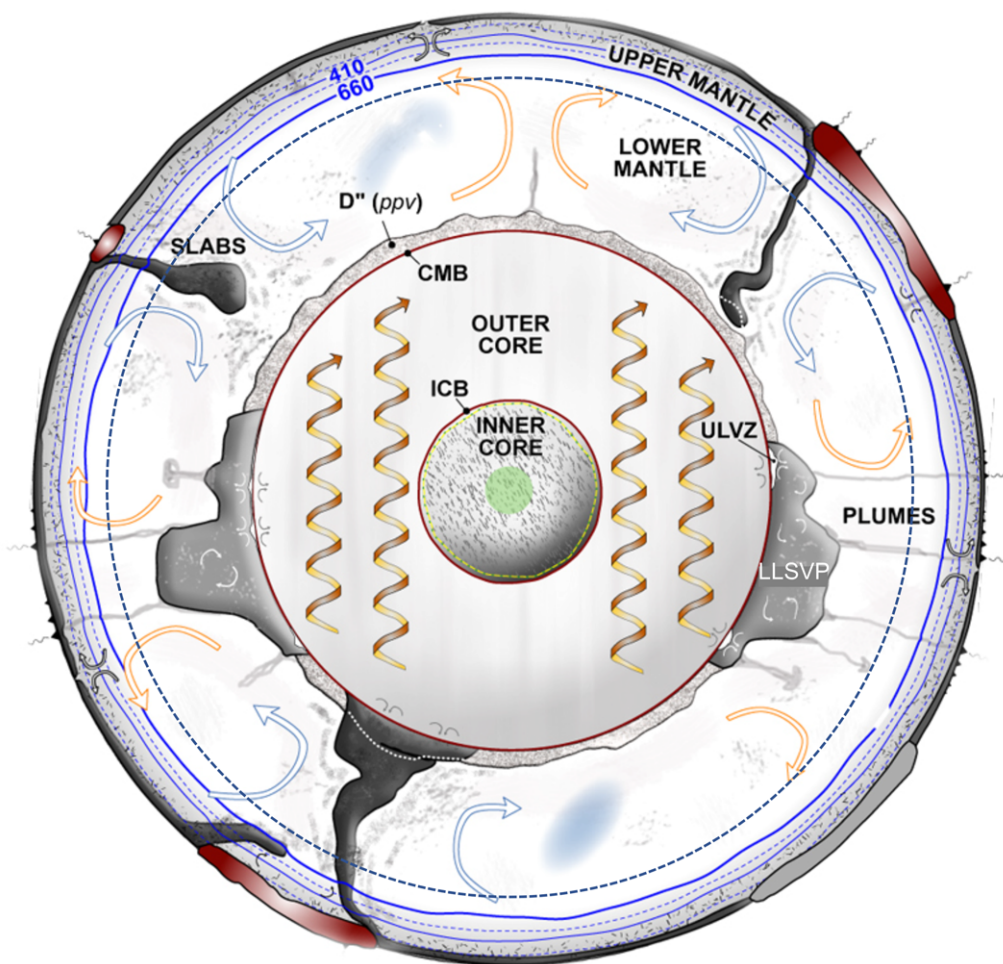
subduction helps to generate plumes through mantle return flow ([Li & Zhong 2009](#)). The configuration of the continental and oceanic plates on the Earth's surface has a strong role in the thermal evolution of the mantle, highlighting the importance of the Wilson cycle in mantle geodynamics. Continental insulation post-supercontinent collision can increase temperatures in the upper mantle (e.g. [Anderson 1982](#); [Coltice \*et al.\* 2007](#)) and the repositioning of subduction zones during the supercontinent cycle may develop deep lower mantle plumes ([Zhong \*et al.\* 2007](#)) ([Fig. 1](#)). The interaction between deep and shallow thermal processes in the mantle during supercontinent formation and dispersal have implications on both local ([Brandl \*et al.\* 2013](#); [Zhang & Li 2018](#)) and global ([Zhong \*et al.\* 2007](#); [Coltice \*et al.\* 2007](#)) scales.

The implementation of numerical models to simulate a supercontinent is a complex endeavour and the results have to be treated with some caution. Although many studies have shown the importance of continental insulation (e.g. [Gurnis 1988](#); [Zhong & Gurnis 1993](#); [Lowman & Jarvis 1993, 1999](#); [Bobrov \*et al.\* 1999](#); [Yoshida \*et al.\* 1999](#); [Phillips & Bunge 2005](#); [Coltice \*et al.\* 2007](#); [Trubitsyn \*et al.\* 2008](#); [Coltice \*et al.\* 2009](#); [Phillips \*et al.\* 2009](#); [Phillips & Coltice 2010](#); [Yoshida 2010](#); [Rolf \*et al.\* 2012](#)), other studies have indicated a lesser impact on mantle dynamics ([Heron & Lowman 2011](#); [Yoshida 2013](#); [Heron & Lowman 2014](#)), alongside cases where continental insulation would, in fact, promote cooling of the mantle ([Lenardic \*et al.\* 2005](#)). In three-dimensional numerical simulations of mantle convection, [Yoshida \(2013\)](#) showed the difficulty in obtaining sub-continental temperatures in excess of sub-oceanic temperatures on timescales relevant to supercontinent episodes for Earth-like Rayleigh numbers, despite the thermal blanket effect of an insulating continent and the formation of sub-continental plumes. [Yoshida \(2013\)](#) indicated that a reversal of mantle motion through the generation of plumes would be sufficient to disperse a supercontinent, despite sub-oceanic and sub-continental temperatures being comparable – highlighting the mechanism of mantle return flow to be important in the supercontinent cycle (e.g. [Zhong \*et al.\* 2007](#); [Li & Zhong 2009](#)). However, the repositioning of subduction to the margins of an oceanic super-plate (with no continental cratonic properties) may also produce deep origin mantle plumes through mantle return flow – allowing the Pacific plate to act as a pseudo-supercontinent ([Heron & Lowman 2010](#)). In addition, the size of the plate has also been indicated to be important in plume generation. [Li & Zhong \(2009\)](#) suggest that the relative strength of plume formation is related to ringed subduction focusing thermal instabilities under the supercontinent, with gradually retreating circum-supercontinent subduction due to

supercontinent dispersal decreasing the intensity of a sub-plate mantle plume.

The processes involved in supercontinent break-up are unlikely to be singular in nature and may be the product of a number of geodynamic phenomenon. It is important to acquire a better understanding of the forces that occur in subduction retreat, continental insulation and plume push (e.g. Wolstencroft & Davies 2017; Zhang *et al.*

2018) to decipher supercontinent break-up. Further geochemical studies on mantle temperatures before and after supercontinent formation would help to improve our understanding of continental insulation (e.g. Brandl *et al.* 2013), in addition to three-dimensional spherical numerical models featuring continental and oceanic plates that move dynamically with mantle convection (e.g. Mallard *et al.* 2016; Coltice *et al.* 2017).



**Fig. 7.** Schematic cross-section of a dynamic Earth with an interconnected crust to core. The mantle is shown with large-scale convective motions (large arrows), primarily driven by the subduction of dense, cold lithosphere (darker outer layer and dark slabs). Whole mantle plumes are most likely to form near or above the hottest deep regions, possibly guided by return flow mechanisms related to subduction and an interaction with large low shear velocity provinces (LLSVPs). The dominant upper mantle phase boundaries near 410 and 660 km depth are deflected by thermal and/or chemical heterogeneities (e.g. slabs and plumes). Other boundaries have also been detected (e.g. the 220, 520 and 1000 km discontinuities, dashed lines). Lower mantle LLSVPs may be reservoirs of incompatible elements and are preferentially located beneath large-scale return flow in the overlying mantle (as discussed in Garnero *et al.* 2016; McNamara 2018). What is clear is that the mantle is a dynamic system with an interior that is changing over time, similar to the Earth's surface. CMB, core–mantle boundary; ULVZs, ultra-low velocity zones; ICB, inner core-boundary. Image modified from Garnero *et al.* (2005).

The Wilson cycle describes the life of an oceanic basin, which culminates in its termination through oceanic subduction. The mechanism of slab-pull, where negatively buoyant lithosphere can propagate into the mantle in a self-sustained manner (e.g. McKenzie, 1977; Gurnis *et al.* 2004), has been indicated to drive the motion of a plate (e.g. Gordon *et al.* 1978; Patriat & Achache 1984), the positioning of spreading centres (e.g. Forsyth & Uyeda 1975; Davies & Richards 1992; Conrad & Lithgow-Bertelloni 2002; Dal Zilio *et al.* 2017) and global-scale mantle flow (e.g. Hager & O'Connell 1981; Becker & O'Connell 2001; Conrad & Lithgow-Bertelloni 2002, 2004). However, an important question remaining on mantle dynamics is its rheological configuration and, indeed, how upwellings and downwellings are impacted by mantle layering on a global or local scale (e.g. Rudolph *et al.* 2015). The life of an ocean below the lithosphere becomes more complicated than simply sinking down, with oceanic subduction having been inferred to interact with various layers within the mantle (e.g. Christensen & Yuen 1984; Fukao & Obayashi 2013) before it reaches the core–mantle boundary and the LLSVPs (e.g. Burke *et al.* 2008; Dziewonowski *et al.* 2010; Steinberger & Torsvik 2012; Conrad *et al.* 2013). Similarly, the interactive nature of plumes with mantle flow and rheological layering can affect their thermal trajectories (Whitehead 1982), alongside plume heads arriving at the mantle lithosphere, to interfere with a Wilson cycle. Kumar *et al.* (2007) suggested that the thinning of the lithosphere from a mantle plume could weaken coupling between the lithosphere and asthenosphere, leading to the increased importance of ridge-push and slab-pull in plate motion. The interaction between mantle flow and continental roots resulting from lithosphere–asthenosphere coupling has been shown to strongly influence plate motion and surface deformation (e.g. Conrad & Lithgow-Bertelloni 2006).

Another key unresolved question is how LLSVPs interact with mantle dynamics over long timescales (e.g. Tackley 1998, 2002, 2011; McNamara 2018), especially the large-scale downwelling related to supercontinent formation. Thermochemical geodynamic models have shown the difficulties in generating positionally stable LLSVPs on timescales >300 myr (e.g. Zhang *et al.* 2010; Tan *et al.* 2011; Li & McNamara 2013). In general, the geodynamic consensus is that downwellings that reach the core–mantle boundary may sweep aside thermochemical LLSVPs (e.g. Tackley 1998; Kellogg *et al.* 1999; Jellinek & Manga 2002; McNamara & Zhong 2005; Bull *et al.* 2009; Lassak *et al.* 2010; Zhang *et al.* 2010; Bower *et al.* 2013; Flament *et al.* 2017) and their current shape may be due to the Earth's subduction history (e.g. Fig. 2) moulding the anomalies

beneath upwelling regions of the Earth (McNamara & Zhong 2005; Bull *et al.* 2009) (e.g. Fig. 1).

It is currently difficult to determine how active LLSVPs are in the generation of plumes, especially if they are shaped by subduction at the surface (e.g. Li & Zhong 2009). Numerical simulations have shown that the specific location of LIPs generated by return flow could be explained, in part, by subduction and mantle viscosity (Davies *et al.* 2012; Heron *et al.* 2015; Li & Zhong 2017). The latter point of mantle viscosity has been raised in the discussion of the persistence of strong silica-enriched domains in the Earth's lower mantle (Ballmer *et al.* 2017). Using geodynamic numerical models, Ballmer *et al.* (2017) showed that the large-scale heterogeneity associated with a 20-fold change in viscosity, such as due to the dominance of intrinsically strong (Mg, Fe)SiO<sub>3</sub>-bridgmanite low-Mg/Si domains, is sufficient to prevent efficient mantle mixing, even on large scales. The stable manifestation of such bridgmanite-enriched ancient mantle structures may reconcile the apparent geographical fixity of deep-rooted mantle upwelling centres (e.g. Torsvik *et al.* 2006; Burke *et al.* 2008; Torsvik *et al.* 2008, 2010).

## Conclusions

The formation of a supercontinent through the closure of an ocean basin has a profound effect on the thermal evolution of the mantle. A large continental plate generates a thermal blanket effect to insulate the upper mantle, resulting in fundamental changes to mantle flow and dynamics (e.g. Anderson 1982; Gurnis 1988; Coltice *et al.* 2007). Continent formation also leads to the cessation of large-scale subduction systems (Fig. 1) and the termination of an oceanic lifecycle, possibly through introversion, extroversion or orthoversion (Fig. 3) (Murphy & Nance 2003; Mitchell *et al.* 2012). The repositioning of subduction (Fig. 2) from this event produces an increase in deep mantle upwellings (Fig. 4) through mantle return flow (Gurnis 1988; Zhong & Gurnis 1993; Lowman & Jarvis 1993; Zhong *et al.* 2007; Li & Zhong 2009; Yoshida 2010; Heron & Lowman 2010; Yoshida 2013). LIPs, believed to be the surface manifestations of these plumes (Courtillot *et al.* 1999), are therefore a consequence of supercontinent formation (e.g. Yale & Carpenter 1998; Ernst *et al.* 2005; Ernst & Bleeker 2010; Sobolev *et al.* 2011) and have been inferred to have a dramatic effect on the lithosphere (Courtillot *et al.* 1999), the atmosphere and oceans (Wignall 2001; Courtillot & Renne 2003; Sobolev *et al.* 2011), and the Earth's magnetic field (Fig. 4) (Larson & Olson 1991; Biggin *et al.* 2012).

A subducted oceanic plate, following the end of its Wilson cycle at the surface, has continued



interactions with the various layers within the mantle (Fukao & Obayashi 2013). The negative buoyancy of the downwelling may have the power to influence the LLSVPs (Fig. 5) at the base of the mantle (Tackley 1998; Kellogg *et al.* 1999; Jellinek & Manga 2002; McNamara & Zhong 2005), which, in turn, may deform the lithosphere once again through plume formation (Fig. 6) (Li & Zhong 2009). A number of geodynamic processes occur during the supercontinent cycle that can lead to continental break-up – the mantle is dynamically linked from crust to core (Fig. 7). Although the Wilson cycle is generally discussed in terms of lithosphere dynamics, the large-scale processes of ocean closure to form a supercontinent can have whole mantle implications through continental insulation, slab-pull and plume formation.

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